One Hundred Plus Years of Wintertime Climate Variability in the Eastern United States*

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ABSTRACT
Interannual anomalies of climate variability in the eastern United States for the past 100+ yr have been studied for their spatial EOF structure, long-term changes, and the covariability with several climate indices: the Southern Oscillation index (SOI), North Pacific index (NPI), and North Atlantic Oscillation (NAO) index. Especially for air temperature, wintertime (December–February) variability is much more pronounced than summertime (June–August). The leading principal component (PC) of wintertime air temperature, which explains 70% of the interannual variance, is significantly correlated with the NAO, while the leading PC of wintertime precipitation correlates with the SOI. The spatial structure of the leading EOFs have a similar spatial character when compared to the correlation between the data and the climate indices, suggesting that the EOFs can be thought of as proxies for mapping the effects of climate indices upon the eastern United States. The effects of the SOI and NPI are generally the same; however, these two climate indices are not independent. The long-term sensitivity of the eastern U.S. climate to the Pacific indices seems only weakly dependent with time, whereas the NAO has grown considerably in importance with time since the beginning of the twentieth century. Surrogate temperature data from New Haven, Connecticut, has been used to extend this 100+ yr analysis back into the previous century, and the apparent long-term trend in the sensitivity to the NAO completely disappeared in the latter part of the nineteenth century. If a measure of potential predictability is the degree to which interannual climate covaries with these climate indices, the recent period (post 1960) may overestimate this predictability based on the long-term changes observed in sensitivity.

1. Introduction
The goal of this endeavor is to understand the spatial patterns and temporal variability of interannual “climate” variability in the eastern half of the United States during the twentieth century. Studies examining climate variability have usually started with the relationship of climate to a given phenomenon, such as the El Niño–Southern Oscillation (ENSO) or the North Atlantic Oscillation (NAO). In particular, Higgins et al. (2000) use tropical Pacific precipitation to define a high-passed, ENSO-like index and a low-passed, North Pacific index–like (NPI) index during the period 1964–93. A third index employed is the Artic Oscillation (AO) index from Thompson and Wallace (1998). The Higgins et al. results are quite promising for potential predictability of U.S. wintertime climate as a significant amount of the interannual signal in temperature and precipitation can be related to these three indices. However, as we shall see, this recent period is one in which climate variability is highly correlated with these climate indices compared to an extended record for the twentieth century as a whole. In this study, we shall start with the dominant spatial patterns and temporal variability in a 100+ yr dataset and then relate different spatial patterns to climate signals. The primary dataset used in this study is based on the National Climatic Data Center (NCDC) divisional temperature and precipitation data (Karl and Knight 1985; NCDC 1994), a summary of monthly minimum and maximum air temperature and precipitation data going back to 1895 and organized by regions within the United States. Various climate indices, their correlation with one another, and their relationships to the air temperature–precipitation data will be examined. In particular, the wintertime period is selected because it exhibits the greatest variability of any season in air temperature, and is most easily tied to various indices, which often serve as metrics of wintertime climate variability. It will be shown that wintertime data in the eastern United States are sensitive to important indices.*

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of climate, and that this sensitivity changes over multidecadal time periods.

2. Seasonal climatologies

We begin with an examination of the seasonally varying surface air temperature and amount of precipitation from a series of selected stations in the eastern United States. Because the spatial scales of variability that we will uncover are large, we have selected 20 representative locations (Fig. 1, Table 1) all having continuous data going back to 1895 and analyzed up through 1999, a total time span of 105 years. The temperature data are available as monthly means of averaged daily minimum and maximum air temperatures. Our analysis focuses on seasonal and interannual variability. Monthly records for the entire period are pooled to first produce a yearly climatology of monthly temperature and precipitation. These climatologies are used to form monthly and then seasonal anomalies. The annual cycle of variability is illustrated (Figs. 2a,b) for three regions: coastal Georgia, coastal Massachusetts, and Missouri. The range of variation of Georgia air temperature is less than Massachusetts, with the former having a yearly minimum (maximum) temperature of 11°С (27°С) and the latter −2°С (22°С). Missouri has the greatest range, having a more continental climate approaching the Georgia maximum in summer and the Massachusetts minimum in winter. For precipitation, Missouri has a wintertime minimum and a spring maximum, Massachusetts has little variation, and Georgia has a large July maximum of nearly 18 cm of rainfall. The monthly anomalies for all 20 sites have been aggregated into 4 seasonal groups: winter (December–February), spring (March–May), summer (June–August), and fall (September–November). The “winter” record for 1900, for example, includes the December data from 1899.

3. Interannual wintertime variability

As interannual variations in the wintertime climate is the largest of the seasons and most tied to large-scale atmospheric and oceanic climate signals, we will concentrate on this season (December–February) for our analyses. We will, however, contrast some of the winter variability with the summer period. We have used empirical orthogonal function (EOF) analysis (Davis 1976) based on the covariance matrices to characterize the dominant modes of interannual variability. For 20 stations, each having equal variance and each being statistically independent, a Monte Carlo simulation revealed that the leading 3 EOFs would account for only about 10%, 9%, and 8% of the total variance, respectively. However, this is clearly not the case (Fig. 3) for either air temperature or precipitation. For the former, the first EOF explains nearly 70% of the total variance, with the second and third EOFs containing 14% and 10%, respectively. For precipitation, the three leading EOFs explain 32%, 21%, and 14% of the variance, respectively. We will therefore concentrate our attention on these leading EOFs, whose spatial structure is given in Fig. 4. In this display a smoother has been used with a search radius of approximately 700 km, before interpolation (a similar approach was used for Fig. 7, below).

The leading temperature EOF T1 is of one sign everywhere, with a broad maximum extending from coastal New England to Georgia and westward to the Mississippi River. The second EOF T2 is positive in the south and negative in the north, while T3 varies from west to east, with largest positive values in the west. For precipitation, the leading EOF P1 is of one sign and
largest in the east with a maximum from Chesapeake Bay to southeast Massachusetts, possibly due to variations of the extratropical storm track and its affect on the coastal region. Second precipitation EOF P2 is positive in the southern United States and negative in the north. Third EOF P3 is large and negative in the south, most prominently over Louisiana, Mississippi, and Alabama and positive in the northeast. We have plotted the leading temperature and precipitation principal components scaled to physical units (Figs. 5, 6) showing a least squares estimate of a trend for each. For contrast, we have also shown the corresponding principal components (PCs) and trends for the summertime period. The leading summertime EOFs of temperature (55% of variance) and precipitation (24% of variance) explain less of their total variance than their wintertime counterparts and less overall variance (especially for temperature). Long-term systematic trends indicate increasing amounts of wintertime precipitation (0.67 cm century$^{-1}$) and increasing air temperatures (0.75°C century$^{-1}$) over the 105-yr record, with smaller or no long-term trends in either summertime precipitation or air temperature. It was noted (Houghton et al. 1996) that in recent decades the increase in air temperature was faster in the daily minimum than the daily maximum temperature. We see here that this same pattern may apply over the eastern United States on a seasonal basis, with larger trends in winter months than in summer. However, given the amount of variance in the two wintertime series, the trends are significantly nonzero only at the 90% confidence level.

On each of the plots in these figures we have also shown the time series of the mean temperature and precipitation anomaly of the 20 stations; in most cases, it is indistinguishable from the leading PC. Since the leading wintertime EOFs represent more of the total variance and that variance is higher in winter than summer, climate variability is predominantly a wintertime phenomenon, especially for air temperature. This is not surprising since the meridional temperature gradient is significantly larger in winter than summer. This can be seen by comparing the temperature difference between Massachusetts and Georgia (Fig. 2) in summer (6°C) versus winter (13°C), the latter having more than a factor of 2 increase in the baroclinicity of the lower atmosphere over the former. The interannual variation in air temperature is substantial. For example, one can see a period in the late 1970s (encompassing the blizzard of 1978) with temperature anomalies of about $-3$°C. The author recalls this period in which one could last walk out onto the waters of Buzzards Bay in Massachusetts, which was covered with sea ice.

a. Relationship to and among climate indices

Since it is of some practical interest to understand the possible sources (and possibly the potential predictability) of climate variability, we consider several different climatic indices. The Southern Oscillation index (SOI) is defined by the normalized pressure difference between Tahiti and Darwin, Australia. The time series

![Fig. 2. Monthly climatology of temperature, (°C, top) and precipitation (cm month$^{-1}$, bottom) for coastal Massachusetts (dashed line), coastal Georgia (solid), and Missouri (dotted).](image)

![Fig. 3. EOF structure for (left) winter temperature and (right) precipitation in terms of the total variance explained by each EOF for their respective interannual time series.](image)
was made available by P. Jones (2000, personal communication; available online at http://www.cru.uea.ac.uk/cru/data/) up through 1997 and afterward from the National Centers for Environmental Prediction (NCEP) Climate Prediction Center; wintertime (December–March) averages were calculated. The Pacific Decadal Oscillation (PDO) index is defined as the leading principal component of North Pacific monthly sea surface temperature variability (poleward of 20°N for the 1900–93 period). Digital values of the PDO index were made available by Mantua et al. (1997; and are available online at http://tao.atmos.washington.edu/pdo/) and we used winter averages (December–March). The NPI (Trenberth and Hurrell 1994) characterizing the strength of the Aleutian low in the North Pacific was averaged over winter months (December–March) and is also available online at (http://www.cgd.ucar.edu:80/cas/cli-mind/np.html). Finally, the NAO index from Hurrell (1995) is the normalized sea level pressure (SLP) difference between Lisbon, Portugal, and Iceland. All of these indices were selected because they had an extended time series throughout the twentieth century. With the exception of the PDO, all of the above indices are also based upon SLP data. Each of these indices has been correlated independently against the wintertime temperature and precipitation data (Table 2). Since each
of the indices are available over a different time span, we chose a common period, beginning in 1900 for this analysis. Therefore, the time span for the display of correlation (100 yr.) for the twentieth century is somewhat different than that of the temperature and precipitation data (105 yr.), that define the various EOFs and PCs. Because the PDO and NPI indices are significantly anticorrelated ($r = -0.57$) and the results that follow are similar (except with a change in sign) for the NPI and PDO, we will not focus on the PDO in our subsequent discussion.

The NAO is positively correlated with air temperatures over the entire region and this spatial structure (Fig. 7, upper left) roughly follows that of EOF1 (Fig. 4, upper left), though with a more southeast U. S. weighting, perhaps due to the influence of the strong NAO signal present in the Sargasso Sea, to the south of the Gulf Stream. The NPI and SOI correlation pattern in Fig. 7 is positive in the south and negative in the north, similar to that of the second air temperature EOF with the SOI correlation better aligned spatially with

![Figure 5](image1.png)  
*Fig. 5. Time variation of the leading PC for EOF1 (dark line) in temperature (top, winter; bottom, summer). The dashed line (all but invisible in this figure but visible in Fig. 6) represents the time series of the numerical mean air temperature anomaly for the 20 sites. The straight line is a least squares fit to the 105-yr trend.*

![Figure 6](image2.png)  
*Fig. 6. As in Fig. 5, but for precipitation, units are cm month$^{-1}$ (the seasonal total anomaly shown by the dashed line is 3 times).*

| Table 2. The correlation between the leading EOFs and the climate indices are given for the periods 1900–99 and 1950–99. We have shown in bold the correlation coefficients that exceed $0.24$ in magnitude for 1900–99. The 99% (95%) confidence limits are $\pm 0.26 (0.20)$ for 1900–99 and $\pm 0.36 (0.28)$ for 1950–99. The same elements shown in bold for 1900–99 are also shown in bold for 1950–99. |
|---|---|---|---|---|---|---|---|---|
| | T1 | T2 | T3 | P1 | P2 | P3 | NAO | SOI | NPI | PDO |
| 1900–99 | | | | | | | | | | |
| T1 | 1 | -0.03 | -0.04 | 0.03 | -0.21 | -0.43 | 0.25 | -0.05 | 0.11 | -0.19 |
| T2 | 1 | -0.01 | -0.36 | -0.38 | -0.16 | 0.18 | 0.44 | 0.37 | -0.25 | |
| T3 | 1 | -0.31 | 0.25 | 0.30 | 0.00 | -0.25 | -0.32 | -0.30 | | |
| P1 | 1 | 0.01 | 0.00 | 0.01 | -0.27 | -0.05 | 0.02 | | | |
| P2 | 1 | 0.01 | -0.12 | -0.49 | -0.45 | 0.35 | | | | |
| P3 | 1 | -0.26 | 0.09 | -0.19 | 0.19 | | | | | |
| NAO | | 1 | 0.03 | 0.16 | -0.06 | | | | | |
| SOI | | | 1 | 0.46 | -0.42 | | | | | |
| NPI | | | | | | 1 | -0.57 | | | |
| PDO | | | | | | | | 1 | | |
| 1950–99 | | | | | | | | | | |
| T1 | 1 | 0.08 | 0.01 | -0.02 | -0.18 | -0.45 | 0.43 | -0.02 | 0.15 | -0.28 |
| T2 | 1 | -0.25 | -0.30 | -0.39 | -0.23 | 0.16 | 0.45 | 0.42 | -0.44 | |
| T3 | 1 | -0.21 | 0.38 | 0.22 | 0.05 | -0.40 | -0.41 | -0.31 | | |
| P1 | 1 | 0.09 | 0.08 | -0.03 | -0.27 | -0.13 | 0.13 | | | |
| P2 | 1 | -0.06 | -0.03 | -0.60 | -0.46 | 0.45 | | | | |
| P3 | 1 | 0.34 | 0.15 | -0.24 | 0.23 | | | | | |
| NAO | | 1 | -0.15 | -0.10 | 0.10 | | | | | |
| SOI | | | 1 | 0.56 | -0.56 | | | | | |
| NPI | | | | | | 1 | -0.73 | | | |
| PDO | | | | | | | | 1 | | |

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the second temperature EOF than the NPI, but with both explaining comparable variance. These correlation maps are similar to the regression plots for hemispheric wintertime temperature study by Hurrell (1996) for a 60-yr period from 1935 to 1994, but we shall soon see that the amount of variance explained can vary substantially over time. For precipitation, the correlation pattern with the NAO is most similar to EOF3 (Fig. 4) with major affects over east Texas and the lower Mississippi River basin. Both the NPI and SOI are similar to precipitation EOF2, with a change of sign, with the SOI influence extending farther to the north into the plains states and along the Carolina coast. This latter probably influences the relatively weak correlation between the SOI and the leading precipitation EOF (EOF1, containing 32% of the variance), but the spatial pattern of this EOF does not match up with any of the correlation patterns as clearly as does EOF2.

The nature of the correlation with the NAO is that most of the region will have warm winter temperatures during high NAO years with wetter than normal winters over the Mississippi River extending into Texas. Posi-
tive phases of the SOI and NPI will produce warmer, dryer winters in the South. Both the NPI and SOI are relatively more important for precipitation overall than the NAO and for temperature their effect is comparable to the NAO in the southeast, but otherwise smaller than the NAO in the eastern United States. This is contrary to many media reports, which often attribute any negative temperature anomalies in New England to El Niño (or La Niña) and all positive anomalies to global warming. Substantial climate variations can arise when two or more indices are in phase or antiphase. For example, a regression against the three indices independently (not shown) would suggest that in the southeastern United States, for example, a wintertime temperature increase of 1°C would result from a positive, two-sigma increase in the NAO index, with a similar temperature increase from a two-sigma increase in the NPI. When these two are in phase and positive (or negative), the effects can be doubled, when antiphase, the effects nearly cancel.

Table 2 summarizes the correlation between the leading EOFs and the climate indices for the periods 1900–99 and 1950–99. We have shown in bold the correlation coefficients that exceed |0.24| in magnitude (the 99% confidence limit is ±0.26) for 1900–99. Those same elements that are bold for 1900–99 are also shown to be bold for 1950–99. In every case in which there was a significant correlation for 1900–99 between the temperature or precipitation “signal” and the climate index or a climate index with another index, the magnitude of the correlation coefficient is either the same (1 case) or increased (14 cases) for 1950–99, suggesting a real increase in importance of eastern U.S. climate associated with these climate indices during the second half of the century. The interrelationships of the climate indices with themselves will be discussed later. Here we note that the NPI and SOI indices are significantly correlated in both periods, in agreement with findings by Trenberth and Hurrell (1994) and Hurrell (1996).

b. Nonstationarity of 100-yr records

The pattern of spatial correlation (as shown in Fig. 7 for the 100-yr period) found for the 1960–99 period (not shown) was examined and found to be similar to that found over the whole period, except for increases in magnitude, especially for the NAO. This suggests that the stability of the EOF pattern is not the issue but rather the changing importance of the different climate indices over time. In order to further examine the change in importance of climate indices over time, 24-yr blocks of data have been selected and their correlation with wintertime modes of temperature and precipitation calculated in a moving average sense, beginning in 1900 and extending until the present. The PCs for temperature and precipitation used have remained constant throughout the period with no recalculation in the 24-yr blocks. Our hypothesis is that the long record best defines the important spatial modes of the data and we seek only to understand how these modes relate to individual climate indices over time. Since the 24-yr records may contain trends (due to longer-term, low-frequency variability), we have calculated a raw correlation and one with both signals first detrended over the 24-yr time block. In Figure 8, temperature PC1 (T1) and precipitation PC3 (P3) are correlated with the NAO index. Over time, the relative importance of the NAO increases steadily for temperature, from levels that are not significant in the first half of the century to significant levels in recent years, with the period from 1960 to the present being the most robust. The precipitation peaks early and late in the record with a distinct minimum in the middle between 1920 and 1940. There is no substantial difference between the raw and detrended correlation in either case. For the NPI and SOI, and PC2 for temperature and precipitation (Figs. 9 and 10), the correlation is much more stable over time with a suggestion of lower values in the middle of the record for the NPI and elevated values for the SOI in recent decades. An interesting feature was found in the SOI analysis (Fig. 10) for precipitation: around 1920 and again in the early 1960s, the correlation of P2 with the SOI was below the 95% significance line. During only these two periods, the correlation (not shown) between the SOI and the leading precipitation mode, P1, became significant. There is general agreement between these figures and Table 2. These long-term variations indicate that the relative sensitivity to different climate indices may vary over time in importance to the eastern U.S. climate as represented by the simple EOF decomposition used here. Correlations using the summertime climate indices for either that year or the previous year...
were examined but found to be less significant than the winter indices for the NAO, SOI, and NPI; periods of low correlation are not due to shifts in sensitivity to a different season of the climate index.

Using a similar approach to that above, the cross correlation of the different climate indices with one another was examined (Fig. 11). The NPI and SOI are significantly correlated over most of the record (again in agreement with Table 2). The period from 1930 to 1945 is one in which all three climate indices are apparently in phase with one another. If one takes the approach of regressing the temperature and precipitation data onto these three different climate indices, one cannot avoid the fact that over some periods, they are correlated; further, they may be correlated over the whole period taken as a basis. This is particularly important for the SOI and NPI as they are significantly correlated over much of the record. Thus, the percent variance explained by the sum of the three regressions on three climate indices is not the same as that expected by summing the $R^2$ values for each separately, because of this non-independence of the indices. A similar result for these two SOI and NPI indices was also found by Hurrell (1996). An alternate approach might be to use the three indices to define an orthogonal index set and use this for the decomposition. It is remarkable that the three indices used by Higgins et al. are only weakly correlated. Unfortunately, they do not extend far enough into the past to be of use for longer-term studies over the twentieth century.

c. An extended look at the NAO influence from 1865

The Hurrell (1995) NAO record of sea level pressure difference between Lisbon and Iceland extends back longer (to 1865) than the atmospheric records used so far. Long time series observations are available at selected locations in the eastern United States all the way back into the eighteenth century. One site, New Haven, in coastal Connecticut, will be used as a surrogate to illustrate some interesting features about the NAO–air temperature correlation, extending the record back until 1865. Here we have combined the New Haven record with the NCDC (Table 1) record from coastal Connecticut, recalculated a seasonal climatology, selected winter months, and correlated this record ($T_{nh}$) with the NAO signal used above with the same 24-yr window, but beginning in 1865 (Fig. 12).

As in Fig. 8, the correlation between the New Haven air temperature ($T_{nh}$) and the NAO is low in 1920 and increases in time, reaching significant levels after 1960. The longer record shows that the initial points on the

FIG. 9. As in Fig. 8, but for the NPI and PC2 for air temperature and precipitation, the latter again multiplied by $-1$ for the presentation.

FIG. 10. As in Fig. 9, but for the SOI and PC2 for temperature and precipitation.

FIG. 11. Moving average correlation among the three climate indices (raw are solid lines and detrended are dashed as in Fig. 8).
earlier curve using PC1 for air temperature were actually the low point in the extended record, with correlation increasing as one moves backward in time to years earlier than 1920. In the late nineteenth century, the correlation is again significant, suggesting that any “trend” in Fig. 8 is really part of a longer period cyclic change. Rather than a long-term increase in sensitivity of air temperature at New Haven to the NAO, we see a substantial decrease in sensitivity in the years 1905–45, when the correlation between the two records is for the most part <0.2. An interesting pattern in the lag-1 autocorrelation of the NAO (especially) and the $T_{ah}$ record as well is a change in character from short, quasi-biennial scales to longer timescales in recent years (Fig. 12, lower panel). The autocorrelation function of the NAO from 1964 to 1999 (not shown) suggests a significant negative lobe at 4-yr lags (corresponding to 8-yr periods). This is in contrast to that in the early part of the record, where the 1-yr lag is negative and significantly greater than zero. These findings are in agreement with Hurrell and van Loon (1997.) who did piecewise spectral analysis of the NAO record over a long period of time and found changes in spectral character of the NAO. Hurrell and Van Loon also found, using an extended NAO record and the wintertime air temperature in Copenhagen, Denmark, that the highest coherence was found in narrow bands that change from quasi-biennial in the early part of the record to having 6–10-yr energy in recent years. These changes do not apparently depend on the value of the NAO since the low-passed NAO record indicates very long timescales of variability with a pair of minima and maxima in the record, although one must acknowledge that the present regime of high NAO may not be over yet. In the present case for Connecticut, it remains a puzzle that a more “tuned,” narrowband NAO signal might be linked to a better correlation with the winter temperature record.

The sea surface temperature (SST) data product presented by Kaplan et al. (1997) and available online up to the present has been used as an additional variable in the analysis. These data are from the winter months (January–March) and represent the average SST in a $10^\circ \times 10^\circ$ square centered on $30^\circ$N, $70^\circ$W, off the east coast of the United States, south of the location of the Gulf Stream (the data location is presented in Fig. 1). This region is one in which the SST and the NAO are positively correlated, the SST signal being one of the three lobes in the SST tripole associated with the NAO (Cayan 1992). The correlation of this signal with the NAO, tracks that of $T_{ah}$ (the two temperature records are significantly correlated) and both reach their lowest levels of sensitivity to the NAO in the same period around 1920. Since both the air and sea temperature correlation with the NAO are decreasing in recent years as the NAO peaks, it is of some interest to speculate on whether the cycle leading to the 1920 minimum will be repeated in the coming decade(s).

4. Discussion

Our initial motivation for starting on this analysis was to understand how wintertime temperatures in the eastern United States varied over time using a 100 yr dataset prepared by NCDC. The spatial patterns of covariability between local measurements and these climate indices closely follow the spatial patterns of the wintertime temperature and precipitation EOFs based on the 100+ years of climate measurements alone, suggesting that the EOFs themselves may independently...
reflect the long-term relationships with various climate indices. Sensitivity to changes in the SOI and NPI were found to be relatively weak during the century. However, the twentieth-century data alone suggested a growing importance over time between the observed eastern U.S. climate and the NAO. An extended analysis using New Haven as a proxy, suggested that this was merely a manifestation of an even longer-term cyclic behavior, with the period of the 1920s one of low sensitivity to the NAO.

As was noted, the Higgins et al. (2000) work was limited to a 30-yr period from 1964 to 1993 and was one in which significant relationships of wintertime U.S. climate were uncovered with variability in the tropical Pacific tied to ENSO, the PDO/NPI, and with the AO, which is essentially the same as the NAO according to Deser (2000). We used a different set of climate indices for our study, in large part because of the extended time series offered by mainly SLP “indices” obtained from local averages (NPI) or point differences in SLP (SOI, NAO). While these point measurements offer an extended time series, they differ from more robust EOF definitions based on a more complete dataset (Deser 2000). Subtle shifts of the centers of action or spatial structure of different climate phenomena may not be reflected in these point indices. Furthermore, our results indicate that restricting one’s attention to recent data will inflate the degree to which different climate indices are estimated to affect eastern U.S. climate. This is most notable for the NAO.

Rogers (1985) has shown that the correlation of the NAO with Oslo, Norway, wintertime temperatures has undergone similar century-long changes with the NAO. In particular, there was a period in the beginning of the twentieth century when the correlation was zero. He attributes this to changes in the zonal location of the Icelandic low for the same values of the zonal flow index (NAO). Thus the pattern of interannual variability in wintertime climate is not completely captured by the NAO index alone. We have also examined long-term temperature records in Oslo and west Greenland, using the same 24-yr running analyses as for New Haven. Oslo is correlated with New Haven and west Greenland is anticorrelated, as we would expect from the thermal anomaly pattern of the NAO as first pointed out by Walker (1924). Oslo has a low point in its sensitivity to the NAO in 1924 (and again in 1960), while west Greenland has its minimum in 1908. This is well within the low-correlation period for New Haven, which exhibits the greatest change in sensitivity of the three sites. Thus, subtle changes in the pattern represented by the NAO index may have more impact on the eastern U.S. winter climate than in northern Europe.

Another potential cause of long-term sensitivity changes may be the changing spectral character of the NAO variability: a whiter spectrum may produce less of a climate signal in the eastern United States than one with energy concentrated in narrow bands, whether they are quasi-biennial or decadal. Thus, an approach that examines the spectral evolution of the coupling, as in Mann and Park (1996), might provide further insight. In that study, the authors noted the degree to which the quasi-biennial temperature variability (mainly in winter) varied significantly over time and had a spatial structure similar the NAO pattern. Thus, changes in both the spatial pattern and the spectral characteristics of the forcing may affect the robustness of the climate response to the NAO.

We have not examined the long-term sensitivity of the SOI or NPI to the eastern U.S. climate extending back more than a century in time as we have above for the NAO. This was in part because the NAO variations over time seem to be larger than the SOI during the twentieth century, although some shift in SOI sensitivity between precipitation EOFs occurred during two time periods in the record. It is also in part because there have been previous studies of long-term sensitivity of tropical SST to ENSO. Elliot and Angell (1988) suggested that long-term changes in the centers of action of the SOI might be at the root of long-term changes in sensitivity of tropical SST to various indices of ENSO. And as noted by Gu and Philander (1995), the varying spectral characteristics of ENSO forcing may also play a significant role in the SST response of the tropical Pacific. The period of low correlation between the NAO and New Haven, centered around 1925, is one in which there appears to be “unusual behavior” in the tropical Pacific, when the Southern Oscillation “faded away” according to Gu and Philander. As noted above, there was nothing remarkable about this period in the sensitivity of eastern U.S. wintertime climate to the SOI, except for the switch in precipitation EOF pattern in the SOI sensitivity already mentioned.

Although the wintertime signals have been the focus of this work, the summertime results shown have indicated that the amplitude of interannual variability in summer temperatures is much smaller than in the winter based on the leading principal component. Furthermore, long-term trends in summertime air temperatures are not seen whereas there is an upward trend in wintertime air temperature and precipitation of about 0.75°C century$^{-1}$ and 0.67 cm century$^{-1}$, respectively, although these are only significantly positive at the 90% level. Finally, we have estimated from the correlation between PC1 of the winter and summer season that there is no significant correlation between the wintertime temperature and that of the preceding summer (0.14), whereas the correlation between the winter and the following summer (0.25) is significant at the 95% level, suggesting that there is some seasonal persistence and therefore some weak predictability of the summer season based on the previous winter.

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